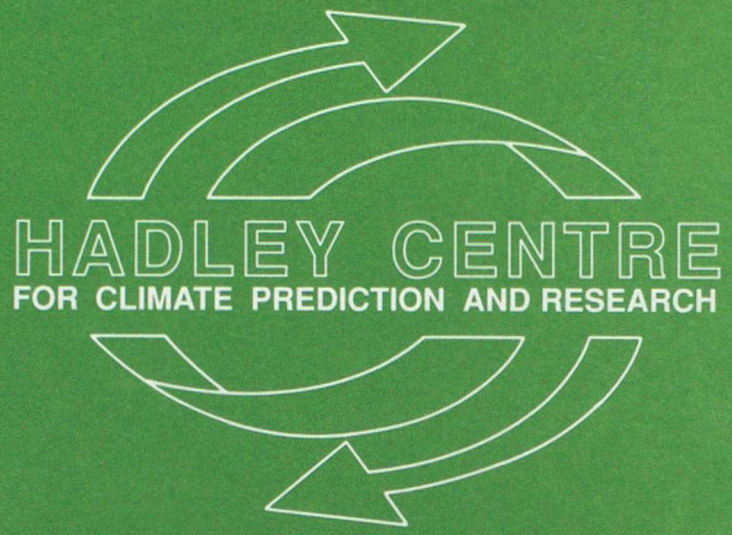

A study of asynchronous coupling using a simple climate model

by

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Hadley Centre
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY

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A STUDY OF ASYNCHRONOUS COUPLING USING A SIMPLE CLIMATE MODEL

by

M K DAVEY

Hadley Centre for Climate Prediction and Research
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY
U.K.

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A study of asynchronous coupling using a simple climate model

Michael K. Davey

Meteorological Office Research Unit
Robert Hooke Institute
Clarendon Laboratory
Parks Road
Oxford OX1 3PU

(Appendix A of Final Report, CEC contract EV4C-0086-UK)

Summary

A very simple linear 3-box thermodynamic climate model is used to develop and test some asynchronous coupling strategies by comparing synchronous and asynchronous experiments. Good agreement is obtained by using a combination of extrapolation of annual averages and interpolation of seasonal perturbations to estimate atmospheric conditions during uncoupled (ocean only) years. Atmospheric forcing with constant and/or seasonally-varying components is considered.

1. Introduction

Very complex general circulation models are required to obtain the best simulations of climatic states and the best estimates of the response of the climate system to changes in conditions such as the concentration of radiatively active gases. Such models contain both atmospheric and oceanic components, with very different response time scales: the atmosphere alone would adjust to new conditions in a few months, whereas the ocean (with its much larger mass and heat capacity) adjusts much more slowly. The timescale for adjustment of the coupled ocean-atmosphere system is thus determined by that of the ocean.

The computational expense of integrating an ocean-atmosphere GCM for centuries in synchronous mode is unfortunately presently prohibitive. ('Synchronous' means that the components are integrated simultaneously with frequent exchange of information about surface fluxes etc.) To obtain balanced climate estimates, either a much simpler, less accurate, model or a different integration strategy must be used.

An alternative integration method is to use asynchronous coupling. The ocean and atmosphere are integrated synchronously for a short time (a year or so, sufficient for the atmosphere and ocean surface layers to adjust, while the deeper ocean changes very little), then the oceanic component is integrated alone for a longer time (several years, so the latest atmospheric conditions are communicated to the deeper ocean), then the cycle of coupled-uncoupled phases is repeated. The atmospheric component is usually more expensive than the oceanic part, because it contains more physical processes and requires shorter timesteps, so the computational savings of asynchronous coupling are large.

Many variations of asynchronous coupling are possible. For example, how many coupled (C) and ocean-only uncoupled (U) years should be used? How should the atmosphere be represented during the

uncoupled phase? One way of addressing such questions is to test the coupling strategy by applying it to simpler, computationally cheap models. In this way synchronous and asynchronous solutions can be compared and unsatisfactory methods can be eliminated. However, there is no guarantee that a method that works well with a simple model will work as well for a much more complex model.

The simplest climate model is a 'box' model, with no dynamics and with each 'box' representing the gross thermodynamic properties of part of the system. Dickinson (1981) used such a model with three boxes and constant external forcing to study asynchronous coupling, and introduced an 'implicitness' parameter to partition atmospheric information used in the uncoupled phase between temperature and heat flux. Roberts (1990) extended that study to investigate the effect of forcing with a seasonal cycle. In this paper further experiments with a similar model are carried out to determine coupling parameters that give accurate results for constant plus seasonal forcing.

Sausen (1988) included stochastic forcing in a 3-box model to represent high frequency variability in the atmosphere. Asynchronous coupling has also been studied by Schneider & Harvey (1986) and Harvey (1986), using somewhat more complex climate models (with more boxes and parameterisations of further physical processes), to find a method with good transient as well as equilibrium behaviour.

2. The '3-box' model

The system used is basically that used by Dickinson (1981) and Roberts (1990): three coupled linear differential equations represent the temperature changes in three boxes that represent the atmosphere, the oceanic mixed layer, and the intermediate ocean. The changes can be regarded as perturbations to an equilibrium climate state, induced

by a thermal forcing anomaly applied to the atmosphere. The equations are

$$C_a \frac{dT_a}{dt} = \lambda_a Q(t) - \lambda_a T_a - \lambda_{am}(T_a - T_m) \quad . \qquad (1)$$

$$C_m \frac{dT_m}{dt} = \lambda_{am}(T_a - T_m) - \lambda_{md}(T_m - T_d) \qquad (2)$$

$$C_d \frac{dT_d}{dt} = \lambda_{md}(T_m - T_d) \qquad (3)$$

where Q(t) is the prescribed energy source. T_a, T_m and T_d are respectively atmospheric, mixed-layer ocean and deeper ocean temperatures. The heat capacities are

$$C_a = 0.45 \ , \quad C_m = 10 \ , \quad C_d = 100 \quad W\,m^{-2}\,year\,K^{-1}$$

and the energy exchange coefficients are

$$\lambda_a = 2.4 \ , \quad \lambda_{am} = 45 \ , \quad \lambda_{md} = 2 \quad W\,m^{-2}\,K^{-1}$$

(These are the same as in Roberts (1990).) Note that time is measured in years. Some relevant timescales are

$$C_a/\lambda_{am} = 0.01 \qquad C_a/\lambda_a = 0.19 \quad years$$
$$C_m/\lambda_{am} = 0.22 \qquad C_m/\lambda_a = 4.17$$
$$C_m/\lambda_{md} = 5 \qquad C_d/\lambda_{md} = 50$$

The fastest timescale (a few days) is for atmospheric adjustment to the ocean mixed-layer temperature. Timescales for radiative equilibration of the atmosphere and for the response of the mixed layer to atmosphere-ocean heat exchange are both about two months. The slowest timescale is for adjustment of the deeper ocean, which

has a large heat capacity.

The forcing is given the simple form

$$Q(t) = Q_0 + Q_1 \sin(2\pi t) \qquad t>0 \tag{4}$$

which represents switch-on of a constant-plus-seasonal energy anomaly. (A gradual increase in Q from zero would be more realistic for a scenario requiring long integration; reproducing the response to a switch-on is a more severe test of the asynchronous scheme.) Temperature anomalies are initially zero. In all cases described below, Q_0 is either 0 or 2.4 Wm^{-2}, and Q_1 is either 0 or 10 Wm^{-2}.

The equations are integrated numerically using leapfrog timestepping with occasional predictor-corrector steps to suppress the computational mode, as is commonly used for more complex climate models. The atmosphere (equ. 1) and ocean (equs. 2 and 3) are treated as separate components. A 0.5 day timestep is needed for the atmosphere, with its short response timescale, whereas a 5 day timestep is used for the ocean. Again, this ratio is typical of more complex models.

3. Synchronous experiments

3.1 constant Q

With Q_0 = 2.4 and Q_1 = 0 the system approaches a steady state with

$$T_a = T_m = T_d = \frac{Q_0}{\lambda_a} = 1\,^oC \tag{5}$$

in the manner shown in Fig. 1. First T_a increases rapidly for a few

days, then the heat flux into the ocean mixed layer raises T_m and slows the rate of increase of T_a. The atmospheric and mixed layer temperatures then increase together on timescale C_m/λ_a (as can be shown analytically), with T_a slightly greater than T_m (Fig. 1a). The heat flux into the deeper ocean then becomes significant, and the timescale of change of all three temperatures shifts to the slowest scale C_d/λ_{dm} during the approach to equilibrium (see Fig. 1b).

3.2 seasonal Q

With $Q_0 = 0$ and $Q_1 = 10$ the linear system must approach a purely oscillatory state with zero annual average. The numerical solution in Fig. 2 shows that T_a and T_m reach equilibrium within a few years, with the amplitude of T_a (about 0.3) larger than that of T_m (about 0.15). (Note that with no ocean the amplitude of the atmospheric response is much larger, about 3.0 .) The response of the deeper ocean (with its large thermal capacity) to seasonal forcing is negligible. The annual average (shown in Fig. 2b for 300 years) remains zero after initial adjustment; the numerical method is stable and accurate.

3.3 constant plus seasonal Q

When $Q_0 = 2.4$ and $Q_1 = 10$ the result is just the linear superposition of cases (a) and (b) above. Fig. 3 shows the transient response for the first 20 years, with the deeper ocean effectively responding only to the annual average of the other temperatures. The annual average of all three temperatures gives a picture almost identical to Fig. 1b.

4. Asynchronous coupling

In the coupled ocean-atmosphere phase equs. 1-3 are all integrated, but in the uncoupled ocean-only phase equ. 1 is omitted, equ. 3 is unchanged, and equ. 2 is altered to

$$C_m \frac{dT_m}{dt} = \lambda_{am}[(1-r)(\hat{T}_a-\hat{T}_m) + r(\hat{T}_a-T_m)] - \lambda_{md}(T_m-T_d) \qquad (6)$$

Here $\hat{T}_a$ and $\hat{T}_m$ are temperatures estimated from values saved during the coupled phase, and r is the implicitness parameter introduced by Dickinson (1981). For r=0 the heat flux into the mixed layer ocean from the atmosphere is determined only by the saved values, and for r=1 the heat flux depends on the saved atmosphere and the current mixed layer temperature. Dickinson found that best results were obtained with a small value for r.

To determine $\hat{T}_a$ and $\hat{T}_m$ we will use annual averages (with $\overline{T}_{-k}$ denoting the average k years before the start of an uncoupled phase), and N 'library' values $TLIB_n$ saved during the last coupled year, as shown in Fig. 4. Here $TLIB_n$ is the average over 1/N years, from time-of-year (n-1)/N to n/N. In all cases described below we choose N=12, so the 'library' values are monthly averages. Other values of N were considered: larger N gives slightly better results but requires more storage (trivial in the present case, but substantial for complex models).

4.1 Constant Q

$Q_0 = 2.4$ is chosen as in section 3.1, so the response should look like the synchronous case in Fig. 1. We first demonstrate the effect of varying the parameter r, using a coupled-uncoupled cycle of 1 year coupled, then 9 years uncoupled (denoted by C,U = 1,9; 1,9; ... .). For these examples $\hat{T}_a$ and $\hat{T}_m$ are estimated by using the monthly averages $TLIB_n$ saved during the coupled year, for the

corresponding months during the uncoupled years. When r=0 (Fig. 5a) the result is clearly unsatisfactory, with large oscillations occurring. Forced by the 'library' heat flux, T_m grows rapidly in the first U phase while T_a simply cycles through its 'library'; T_a then increases rapidly in the second C phase, but does not catch up with T_m, leading to a stored heat flux that substantially reduces T_m during the second U phase, etc. The case r=1 (Fig. 5b) is also unsatisfactory, because T_m is closely coupled to the T_a 'library', which is a bad estimate of true T_a behaviour (Fig. 1). Real increase in T_m only occurs in the C phases. (Note that the apparent seasonal cycle in the U phases is artificial: there is no seasonal forcing, and the cycle in T_a only appears due to the annual repetition of the monthly 'library' values.)

Some compromise between these two extremes is needed. From an analytic 2-box model (no deeper ocean) Dickinson (1981) estimated that $r = \lambda_a/(\lambda_a + \lambda_{am})$ is the required value. For the parameters of the present model this value is about 0.05, giving the considerably improved behaviour shown in Fig. 5c. (See also Fig. 3 in Roberts (1990) for the effect of other values of r.) This value r=0.05 is used in all subsequent cases.

The result is still poor in the first few years of rapid change. Further, saving the 'library' with the coupled phase lasting only one year means that the 'library' contains transient adjustments caused by the fresh coupling. Thus an initial 5 coupled years were used in the next case, with 2 C years and 8 U years subsequently (C,U =5,8; 2,8; 2,8;...). Fig. 6 shows that the behaviour in the first 30 years is considerably improved, though there is little difference thereafter when changes during one C,U cycle are relatively small.

Simply repeating the seasonal 'library' during the uncoupled phase is clearly not a good representation of true interannual behaviour. The next step was to try a different estimate based on

linear extrapolation of the saved annual averages $\bar{T}_{-k}$. Define

$$\tilde{T}(t) = T_0 + \alpha \frac{(\bar{T}_{-1} - \bar{T}_{-2})}{\Delta t} (t - t_0) \qquad (7a)$$

$$T_0 = \bar{T}_{-1} + \frac{1}{2} \frac{(\bar{T}_{-1} - \bar{T}_{-2})}{\Delta t} \qquad (7b)$$

where T_0 is the estimated temperature at the time t_0 at the end of the coupled phase, α is a factor controlling the slope of the extrapolated temperature during the uncoupled phase, and $\Delta t = 1\,year$ (see Fig. 7. For the first extrapolated case shown in Fig. 8 the cycle is again C,U = 5,8; 2,8; 2,8; ... (for comparison with Fig. 6), but now $\tilde{T}_a = \bar{T}_a$ and $\tilde{T}_m = \bar{T}_m$, with α=0.5 . This choice of α is a compromise: T_a is overestimated in the first uncoupled phase but underestimated in later phases. The ocean temperatures in Figs. 6 and 8 are very similar; only the T_a patterns differ markedly. (Remember that with r=0.05 the ocean mixed layer is forced mainly by $\tilde{T}_a - \tilde{T}_m$, rather than $\tilde{T}_a - T_m$.)

For the next extrapolated case we choose (after trying several alternatives) C,U = 10,17; 3,17; 3,17; ... , and α=0.7 . The initial rapid change is the most difficult stage to reproduce, so 10 initial coupled years are used. The extrapolation is based on two coupled years, so we use at least three coupled years to allow for initial adjustment after coupling. To maintain a high ratio of U to C years, 17 U years are used. Fig. 9 shows results for 300 years: the T_m and T_d curves are almost identical to the synchronous case (Fig. 1), and apart from the first U phase T_a is also very close to the synchronous response.

Attempts were made to improve $\tilde{T}$ by using more than two saved annual averages (ie to include curvature information), but the

results were not robust because small fluctuations in the saved values could produce large errors in the long uncoupled phase. Small improvement could be made by by reducing α as equilibrium is approached, when adjustment due to renewed coupling starts to exceed long-term trends.

4.2 Seasonal Q

Having established a suitable algorithm for the response to constant Q alone, the next task is to consider seasonal effects in isolation, making use of the seasonal 'library' during the uncoupled phase. Initial experiments simply used the 'library' values $TLIB_n$ for the corresponding time of year in the uncoupled phase, but with N=12 values (monthly averages) a significant phase error was thus introduced. To overcome this problem, the 'library' values were linearly interpolated, taking the n^{th} value as that for time-of-year $(n-\frac{1}{2})/N$ (the central time in the averaging period).

With C,U = 5,8; 2,8; 2,8; ... and r=0.05 the temperature signals are very similar to the those in the synchronous case shown in Fig. 2. The difference (synchronous - asynchronous) in T_m is given in Fig. 10 for the first 20 years. The difference is of course zero during the initial coupled phase; it then oscillates during the uncoupled phases with an amplitude less than about one tenth of the full T_m signal, with smaller irregular fluctuations in succeeding coupled phases. (The same pattern continued for 300 years of integration, with a very similar pattern for the T_s difference also.) Note that the difference is in-phase with the full signal when uncoupled: the difference arises mainly because the asynchronous T_m has slightly smaller magnitude than the synchronous T_m. This difference is due to the smoothing of peaks when the monthly average 'library' values are saved during the coupled phase; by using more 'library' values per year the difference can be reduced further.

This purely seasonal case was repeated with r=0 and r=1. In contrast to the response to constant Q (section 4.1), results were not sensitive to the choice of r. The implicitness parameter mainly serves to ensure that underlying long-term trends are correctly estimated during uncoupled periods.

4.3 Constant plus seasonal Q

The techniques established above can be combined to calculate $\bar{T}$ for the uncoupled phases. The interannual trend is removed from the N seasonal 'library' values by defining, for time-of-year $t_n = (n-\frac{1}{2})/N$,

$$TLIB'_n = TLIB_n - \left[\frac{\bar{T}_{-1}+\bar{T}_{-2}}{2} + \frac{\bar{T}_{-1}-\bar{T}_{-2}}{\Delta t}\, t_n\right] \qquad (8)$$

Perturbations T' are obtained by linearly interpolating this perturbation 'library', and these are added to the extrapolated interannual trend $\bar{T}$ to obtain

$$\hat{T} = \bar{T} + T' \qquad (9)$$

As in section 3.3, Q_0 = 2.4 and Q_1 = 10 is used for the example. With C,U = 10,17; 3,17; 3,17; ... , r=0.05 and α=0.7 the asynchronous results agree very well with the synchronous case at all times. As a typical sample, Fig. 11 shows the synchronous and asynchronous T_m signals for years 285 to 295 (years 288 to 290 are coupled), when the system has almost reached equilibrium: phase agreement is excellent, and the amplitudes are also very close. The T_m difference (synchronous-asynchronous) for the first 50 years can be seen in Fig. 12a. Like the purely seasonal case (section 4.2) the difference is small and in phase with the full signal, even during this period of relatively rapid interannual change. The annually-

averaged errors are an order of magnitude smaller still, with no long-term trend.

The deeper water is insensitive to the fluctuating errors in mixed-layer temperature. Fig. 12b shows the T_d difference for 300 years: the error is indistinguishable from the accuracy (+-0.001) of the values saved for comparison.

The T_a difference is also of interest (Fig. 12c). Small oscillations are seen during the uncoupled phase (when T_a is just the estimated value $\hat{T}_a = \bar{T}_a + T'_a$), again due to the amplitude underestimate by monthly averages. T_a is generally estimated well however; this feature may be important if the forcing Q depended on atmospheric conditions on the C,U timescale.

5. Discussion

Complicated climate models may be integrated relatively economically by using asynchronous coupling. A very simple 3-box thermodynamic climate model with constant and seasonal forcing has been used to investigate this technique. It is found that the results of synchronous integrations can be reproduced accurately, by using a mixture of extrapolation of annual averages and interpolation of seasonal perturbations to estimate atmospheric temperature and surface heat flux during the uncoupled (ocean-only) phases of an asynchronous integration. Best results were obtained when at least three coupled years were used between each (much longer) uncoupled phase: in saving values for use in the uncoupled phase the first coupled year was discarded (because it contains some rapid adjustment induced by the new coupling) and then two annual averages were required for extrapolation.

The forcing was switched on suddenly, causing relatively rapid adjustment of the system for the first few years. During the asynchronous integration the ocean and atmosphere were initially

coupled for several years to best represent this behaviour. (An alternative would have been to use a sequence of short coupled and uncoupled phases.) For anomalous climate forcing that arose gradually this initial coupled sequence would not be required.

There was very good agreement in the phase of the synchronous and asynchronous seasonal cycles. The main residual error was due to underestimation of the amplitude of the seasonal cycle, caused by the use of monthly averages in the 'library' values. This slight error could be corrected by rescaling: in a more complex system the error would most likely be lost in other noise however.

The method described should work well for situations where the annual average and seasonal cycle of the external forcing vary slowly compared to the cycle of coupled and uncoupled phases. The method should also work for more realistic systems where the forcing anomaly depends on the internal variables, because the atmospheric changes during uncoupled phases are estimated well on a monthly timescale.

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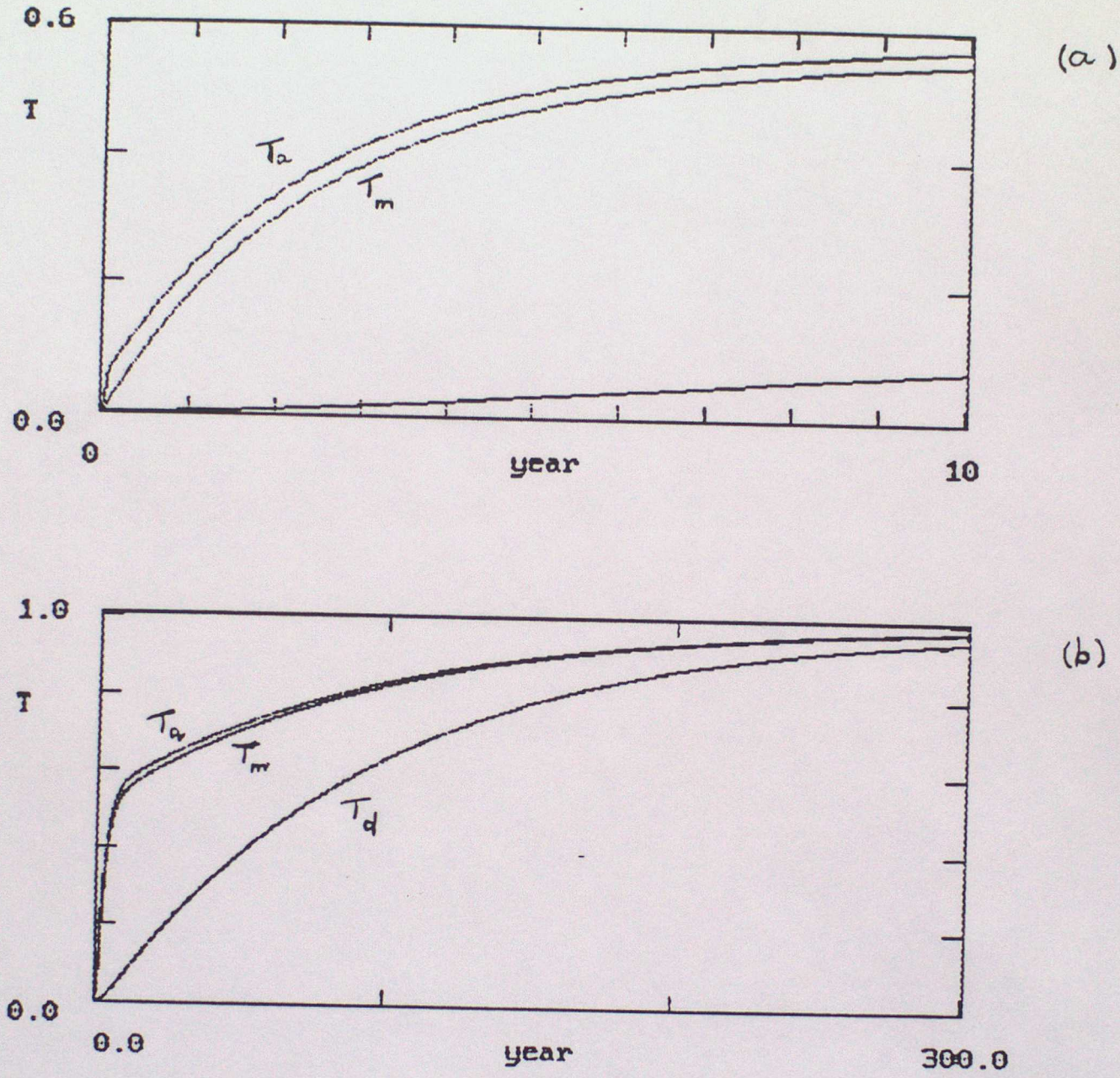

Fig. 1: Synchronous integration with constant forcing Q=2.4
(a) Detail of the first ten years.
(b) Gradual approach to equilibrium as the heat flux into the deeper ocean slows the initial more rapid response of the atmosphere and the mixed-layer ocean.

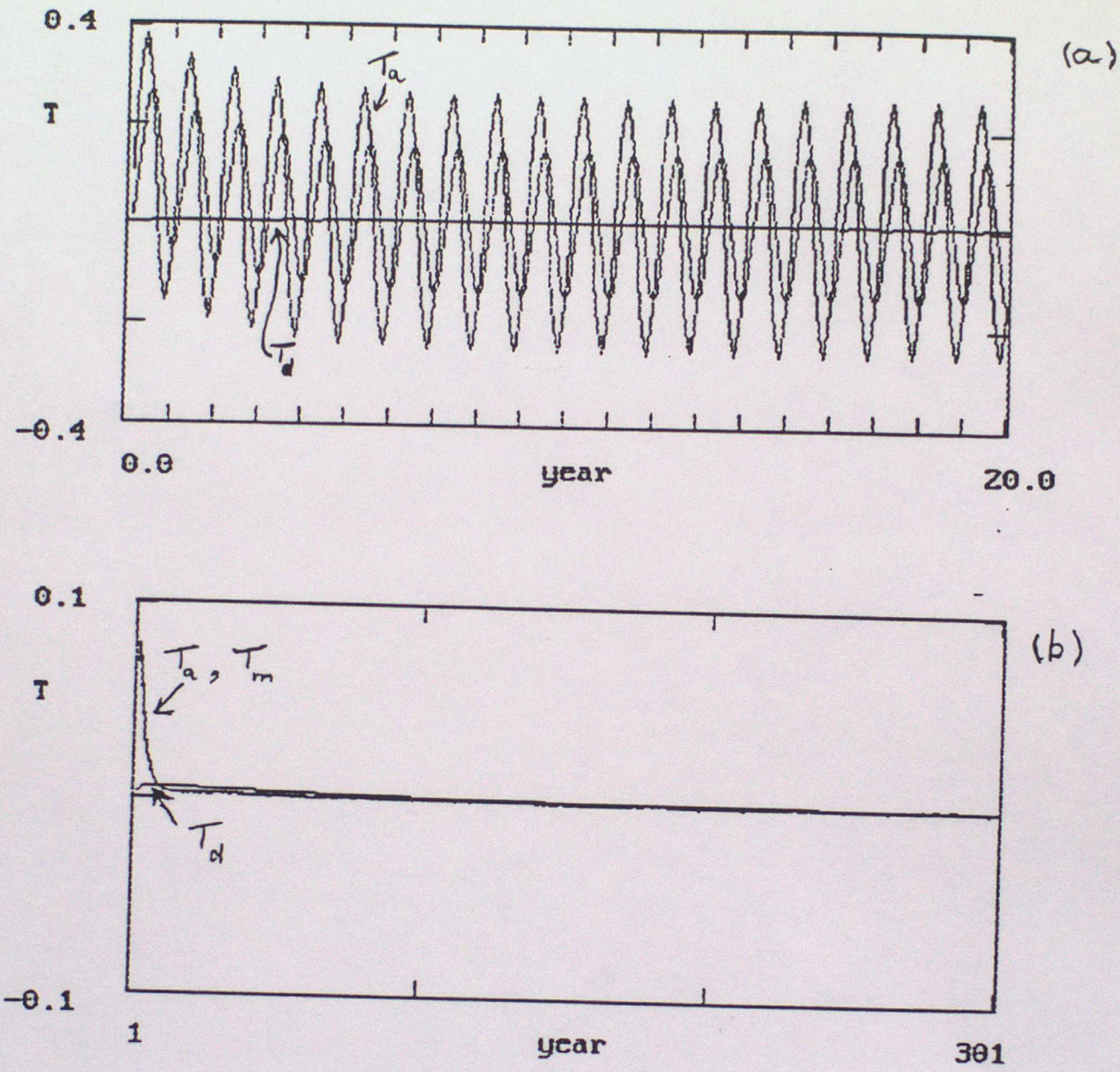

Fig. 2: Synchronous integration with seasonal forcing $Q = 10 \sin(2\pi t)$ switched on at time zero.
(a) The first 20 years, during which the system settles into a stable seasonal cycle. The deeper ocean is almost unaffected.
(b) Annual mean temperatures for 300 years.

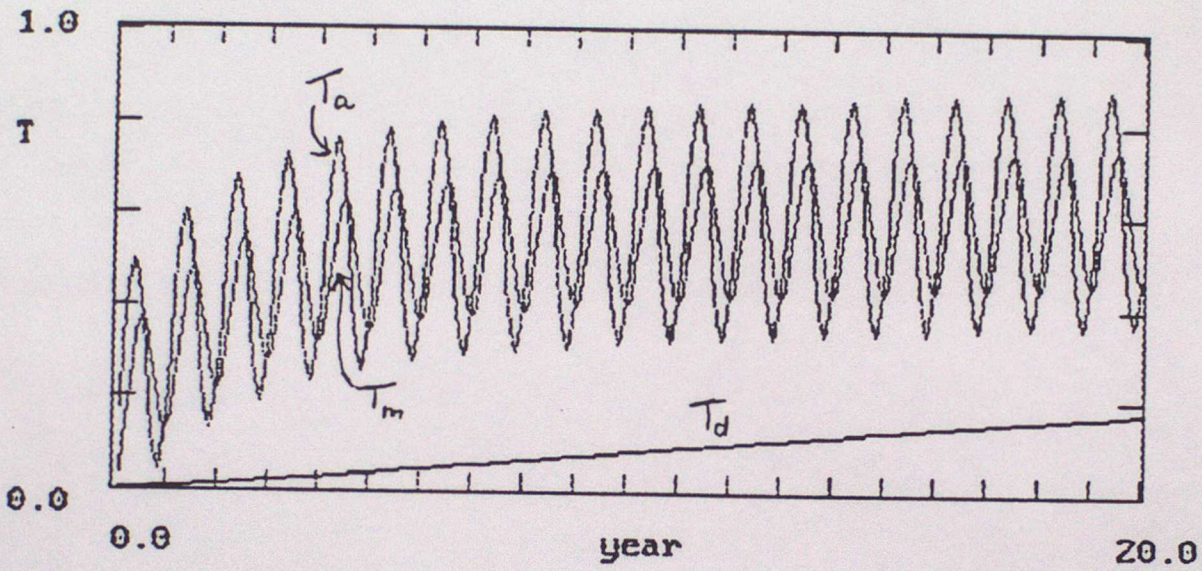

Fig. 3: Synchronous integration with $Q = 2.4 + 10\sin(2\pi t)$ switched on at time zero.

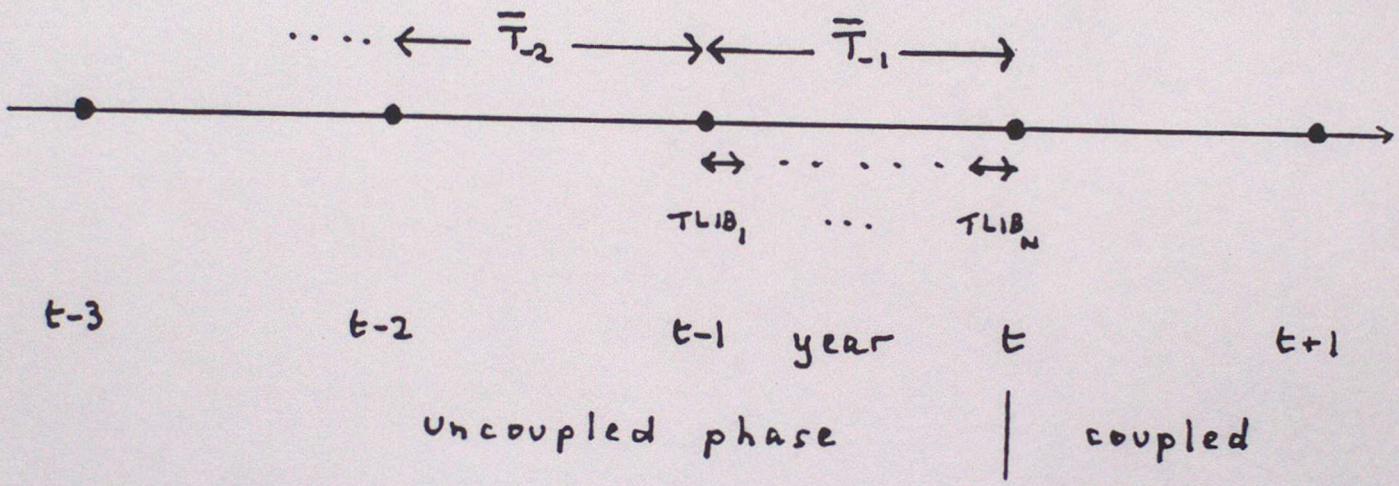

Fig. 4: Annual averages $\bar{T}_{-k}$ and intra-annual average 'library' values $TLIB_n$ are saved during the coupled phase for use during the subsequent uncoupled phase.

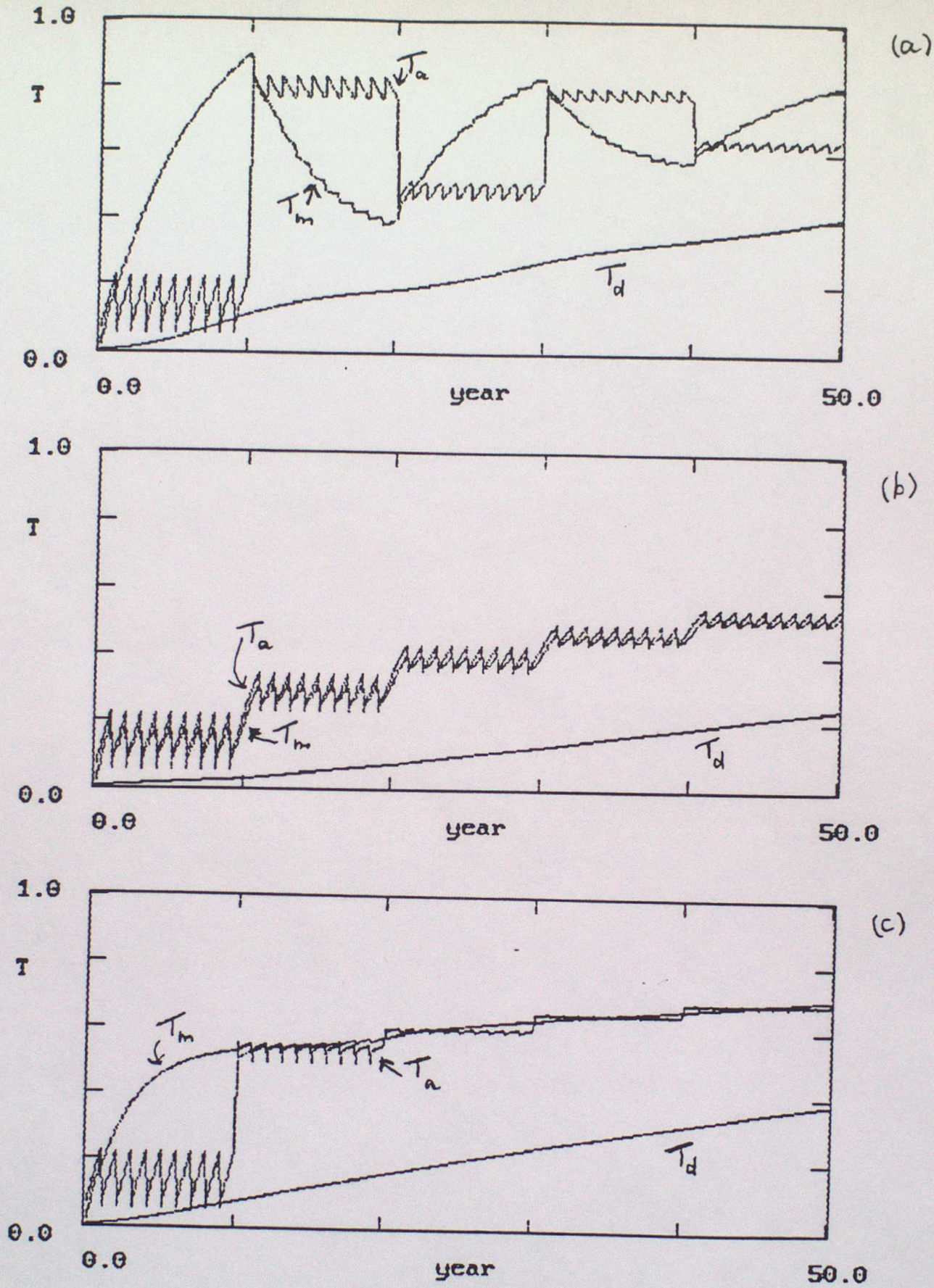

Fig. 5: Asynchronous integration with constant forcing Q=2.4 from time zero. Coupling cycle C,U = 1,9; 1,9; ... . The monthly average 'library' $\overline{TLIB}_n$ is used in ocean-only uncoupled phases without interpolation, and with no interannual extrapolation. During uncoupled phases the estimate $T_a = \hat{T}_a$ is plotted.
(a) implicitness parameter r=0
(b) r=1
(c) r=0.05

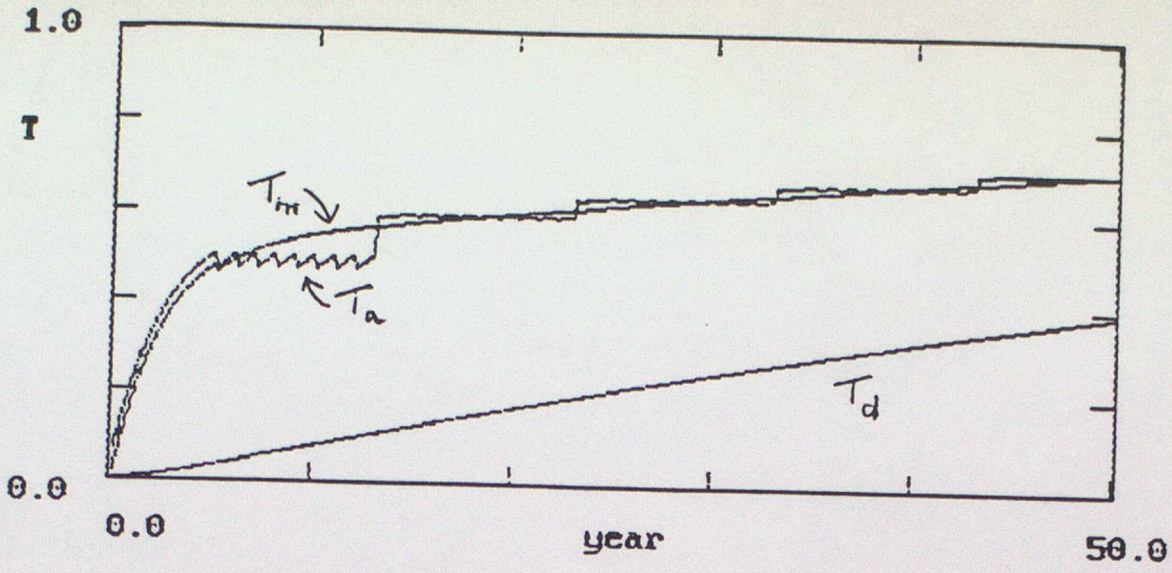

Fig. 6: Asynchronous integration with constant forcing Q=2.4 from time zero. Coupling cycle C,U = 5,8; 2,8; 2,8; ... , implicitness parameter r=0.05. The monthly average 'library' $TLIB_n$ is used in ocean-only uncoupled phases without interpolation, and with no interannual extrapolation. During uncoupled phases the estimate $T_a = \hat{T}_a$ is plotted.

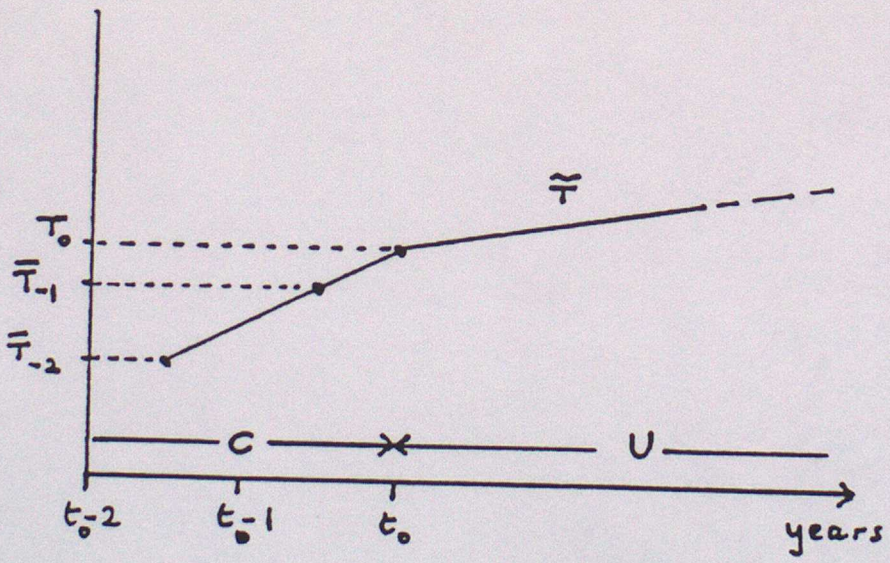

Fig. 7: Extrapolation of annual average temperatures $\bar{T}_{-k}$ saved during coupled years.

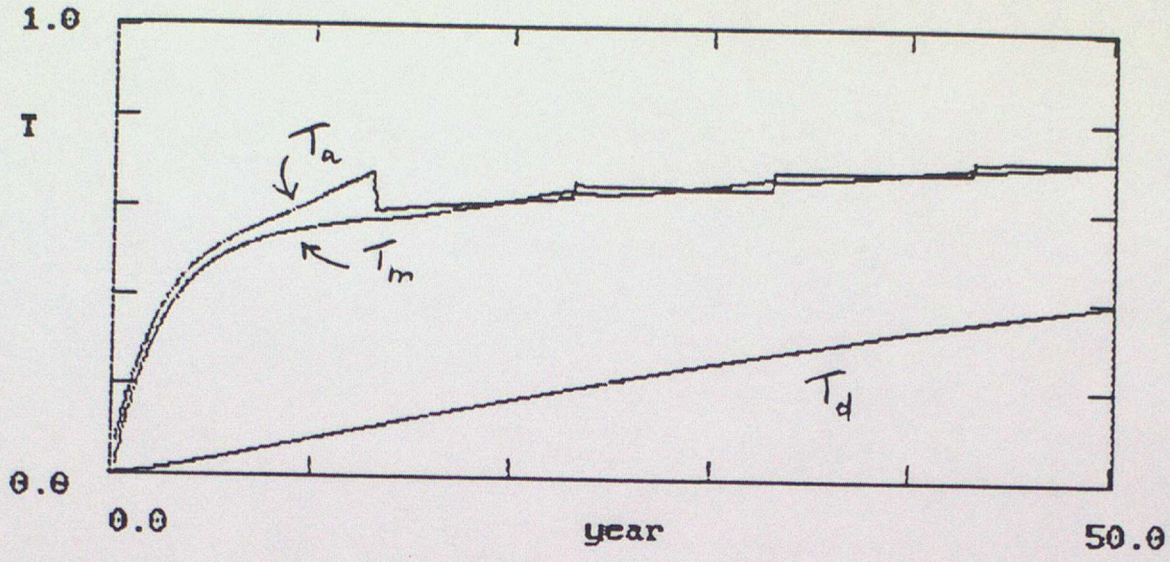

Fig. 8: Asynchronous integration with constant forcing Q=2.4 from time zero. Coupling cycle C,U = 5,8; 2,8; 2,8; ... , implicitness parameter r=0.05. During ocean-only uncoupled phases the extrapolated annual average $\bar{T}$ is used, with slope parameter α=0.5 .

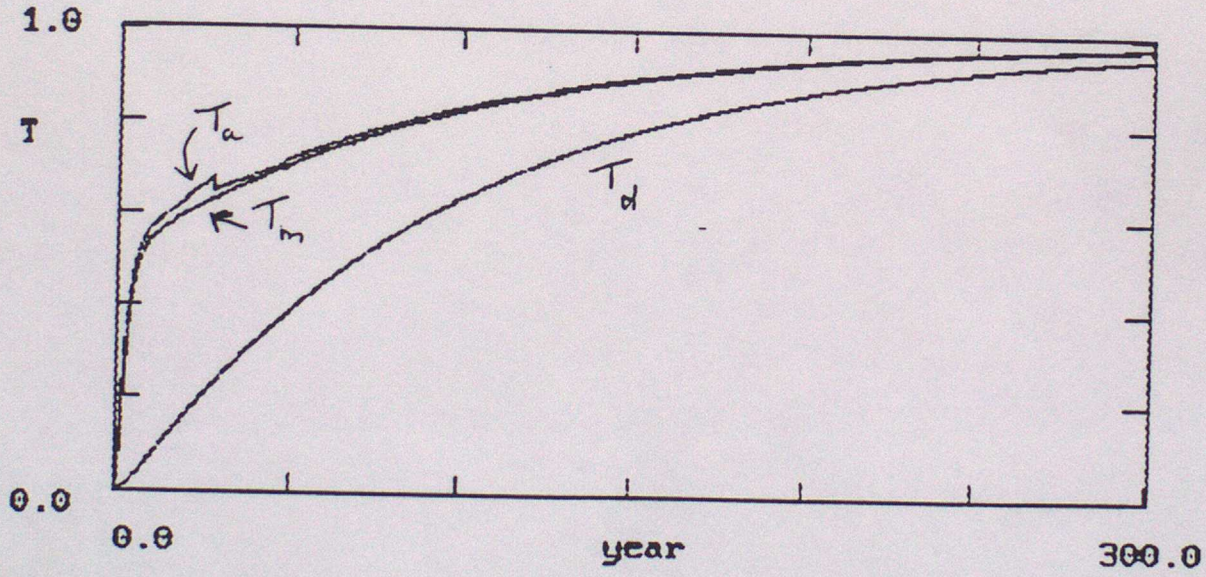

Fig. 9: Asynchronous integration with constant forcing Q=2.4 from time zero. Coupling cycle C,U = 10,17; 3,17; 3,17; ... , implicitness parameter r=0.05. During ocean-only uncoupled phases the extrapolated annual average $\bar{T}$ is used, with slope parameter α=0.7 .

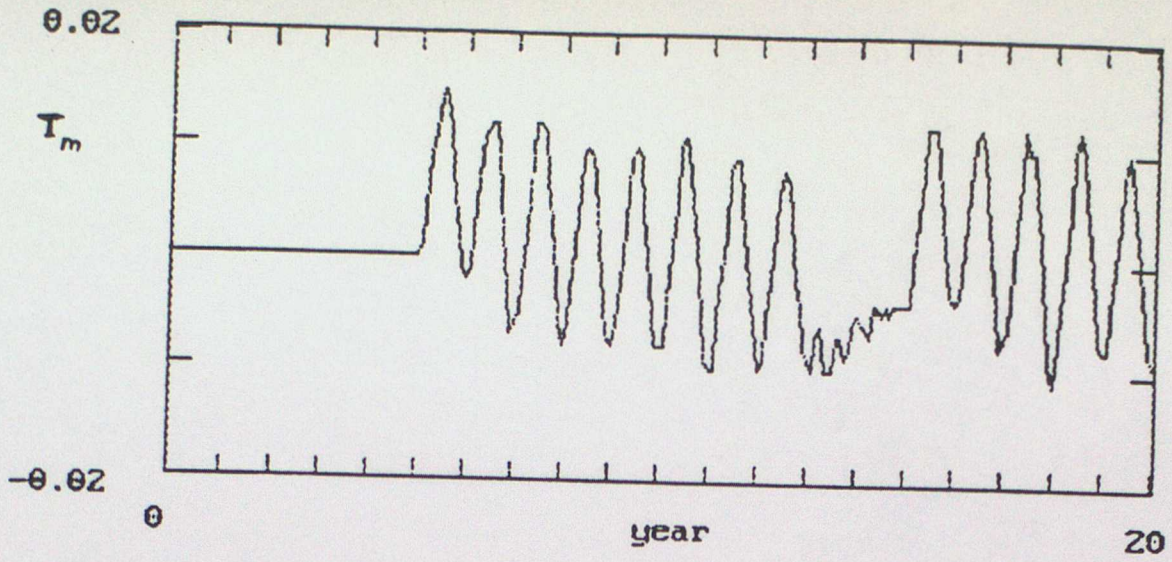

Fig. 10: The difference (synchronous - asynchronous) in ocean mixed layer temperature, in response to seasonal forcing $Q = 10\sin(2\pi t)$ switched on at time zero. For the asynchronous integration C,U = 5,8; 2,8; 2,8; ... , and implicitness parameter r=0.05. During the ocean-only uncoupled phases monthly average values $TLIB_n$ are used with interpolation.

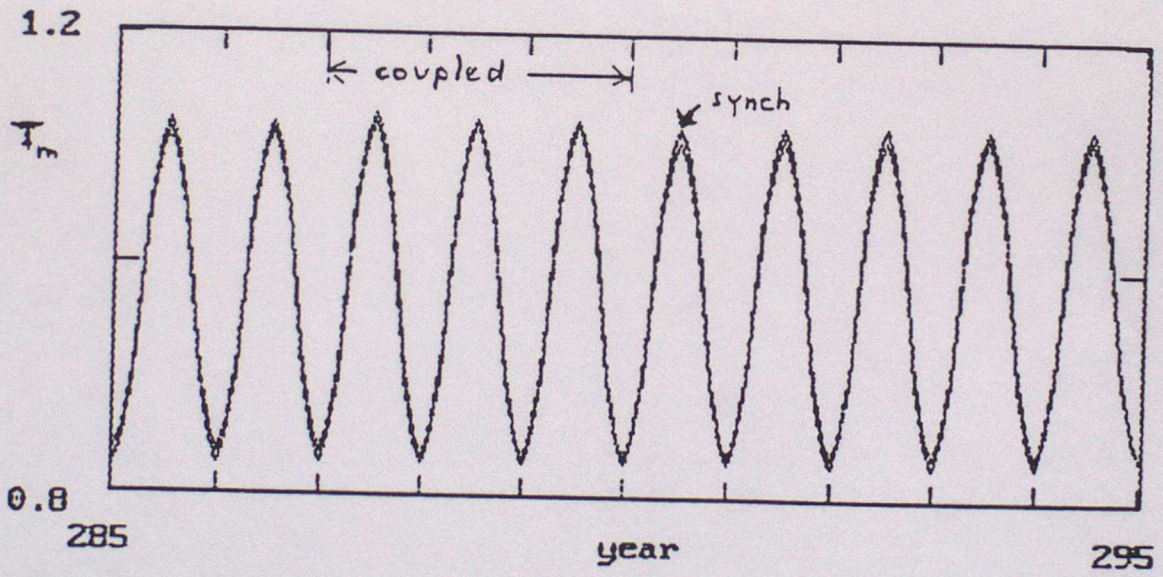

Fig. 11: Synchronous and asynchronous ocean mixed layer temperatures, in response to forcing $Q = 2.4 + 10\sin(2\pi t)$ switched on at time zero. For the asynchronous integration C,U = 10,17; 3,17; 3,17; ... , and implicitness parameter r=0.05. During the ocean-only uncoupled phases extrapolated annual averages (with slope parameter α=0.7) are added to interpolated monthly average perturbations $TLIB'_n$.

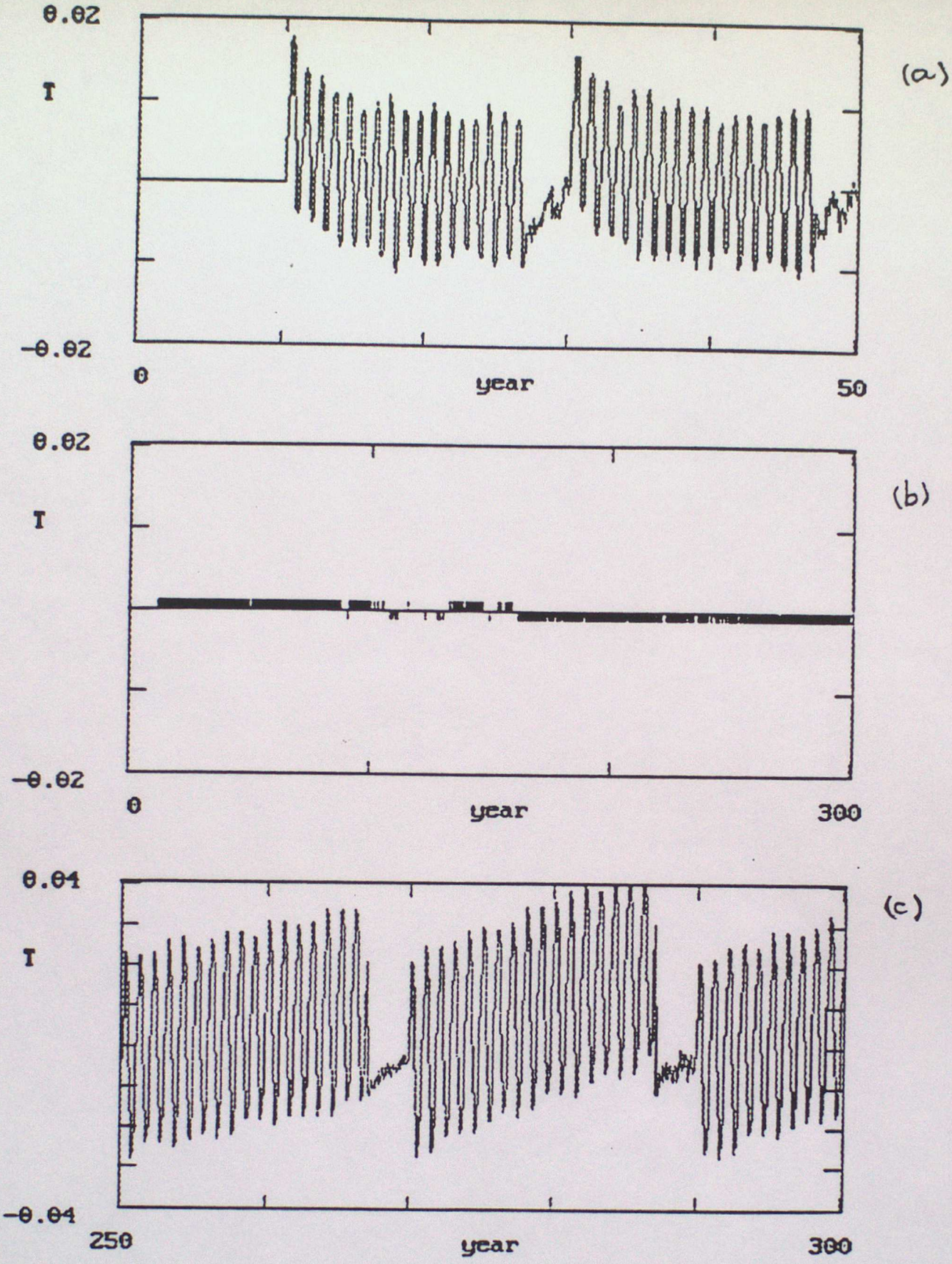

Fig. 12: Temperature differences (synchronous - asynchronous) for forcing $Q = 2.4 + 10\sin(2\pi t)$ switched on at time zero. For the asynchronous integration C,U = 10,17; 3,17; 3,17; ... , and implicitness parameter r=0.05. During the ocean-only uncoupled phases extrapolated annual averages (with slope parameter α=0.7) are added to interpolated monthly average perturbations $TLIB_n'$.
(a) Ocean mixed layer temperature, year 0 to 20.
(b) Deeper ocean temperature, year 0 to 300.
(c) Atmospheric temperature, year 250 to 300. During uncoupled phases $T_a = \hat{T}_a$ is plotted.

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